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Extraterrestrial soils — the lunar experience

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Introduction

Soils form on planetary surfaces, that is, at the interface between the planet and its interplanetary environment, in response to the release of internal planetary (endogenic) energy and the flux of extraplanetary (exogenic) energy (Figure 3.1). It may seem simplistic to make such a statement but it is essential to divest ourselves of an earthbound conceptual framework if we are to place terrestrial soils into the larger framework of planetary processes. On planets such as the earth the atmosphere and hydrosphere act as intermediaries, redistributing the available energy flux in a complex way across the whole planetary surface. As a consequence, extraterrestrial energy fluxes, for the most part solar energy, are redistributed by mechanisms such as wind, waves or stream flow. On smaller planetary bodies with proportionately smaller gravitational fields, a planet's ability to hold an atmosphere is reduced such that on a body the size of the moon, for all practical purposes there is no atmosphere or hydrosphere. This absence is critical in determining the path that soil formation takes. Without air and water many of the things we take for granted in the terrestrial environment simply disappear; there is no rainfall, no runoff and no stream erosion and consequently there are no clays, no hydrated minerals, no water table and so on. We literally find ourselves in another world.

The surface of the moon and other smaller planetary bodies is exposed directly to the interplanetary environment. Without fluid intermediaries solar energy, the dominant energy source in the terrestrial setting, becomes largely ineffective as a source of erosive soil-forming energy because it is simply re-radiated back into space from the planet's airless surface. Conversely, the meteorite flux, a feeble exogenic energy source in the terrestrial setting, becomes, by default, the dominant source of erosional and transportational energy available at the surface of the planet. The only other soil forming processes that may affect soil development are mass wasting, which aids erosion on steeper slopes (Swann et al., 1972), and electrostatic transport, which occurs along the planet's terminator (Criswell, 1972).

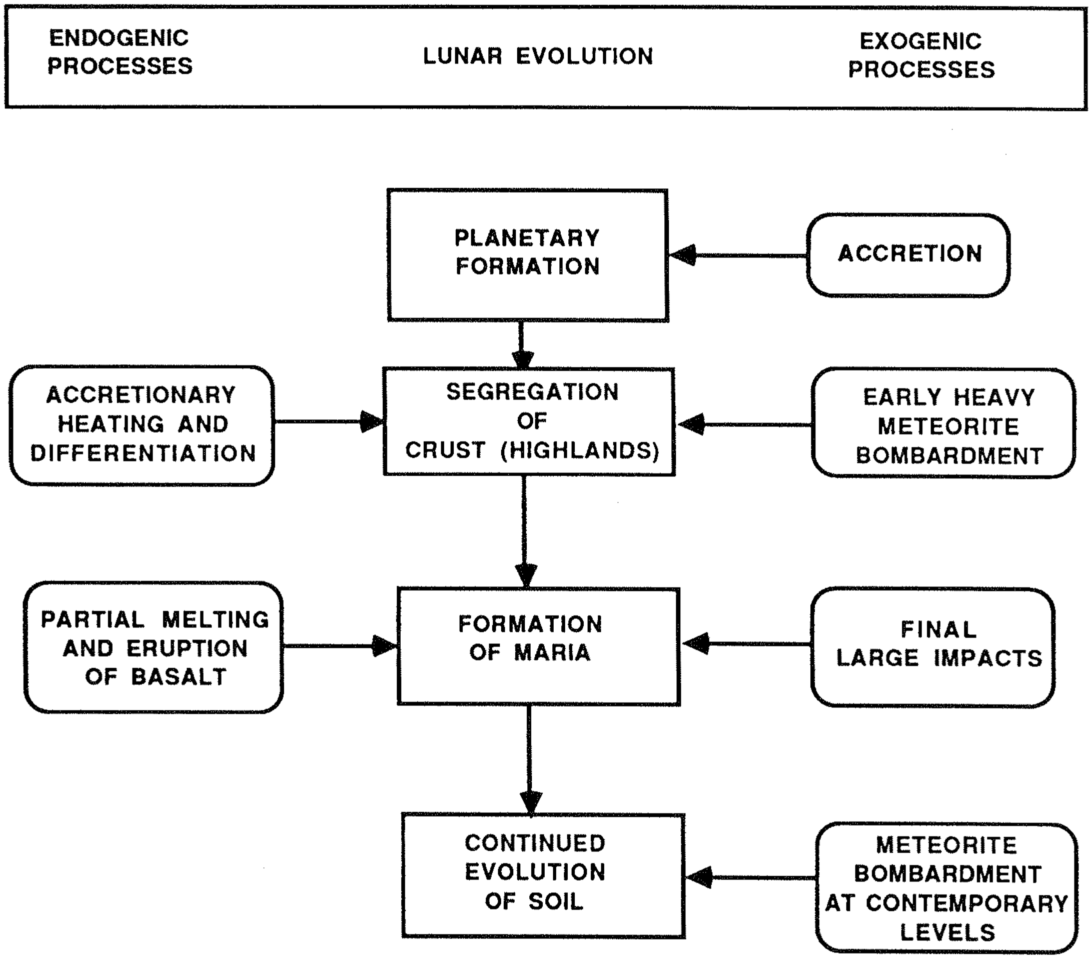

Fig. 3.1. Simplified schematic relationship between planetary processes and soil formation on the moon (after Veverka et al., 1986). The soil is a product of the composition of the moon's surface and its geologic history which is a product of both endogenic and exogenic processes.

Electrostatic transport may be important in transporting finer material less than 10 μm in diameter.

The moon

When viewed from the earth the moon's surface is obviously divided, using albedo, into maria and highlands (terra; Figure 3.2). The highland areas are consistently rougher and more intensely cratered than the maria suggesting that there is a considerable difference in age. These simple observations, first made by Galileo in 1610, provided the first direct information that the moon, like the earth and other planets, has had a complex evolution. The moon was probably formed 4.65×10^9 years ago. The large-scale chemical differentiation of the moon began almost immediately and was complete shortly after 3.85×10^9 years ago. By this time the lunar crust, which is of most interest to us here, had formed and become cold and rigid; the moon, to all intents, had ceased to evolve and had become a dead planet (Figure

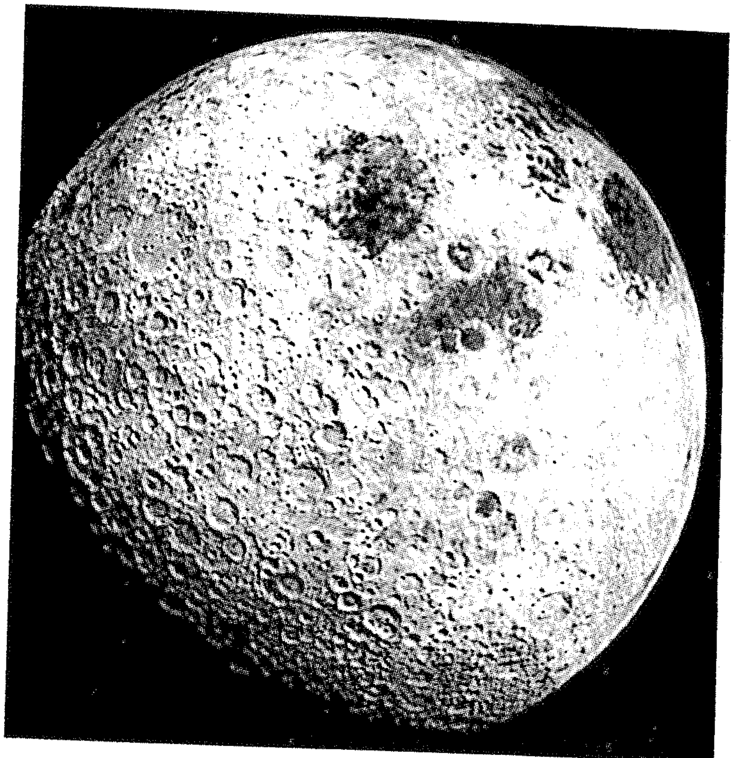

Fig. 3.2. A view of the intensely cratered lunar farside and part of the nearside including Mare Crisium. Note the lack of large craters on the younger mare surface indicating a considerable age difference (NASA AS-16-3028).

3.2). The two-fold subdivision of the lunar surface thus reflects its evolution. The highlands, which on average are 3–4 km higher than the maria, are underlain by much altered lunar crust, whilst the maria were formed by lava flows in the final stage of the planet's evolution. Highland lithologies are, in general, anorthositic in composition whilst the lavas of the maria are basaltic. This compositional difference determines, in large part, the composition of the soil formed on the bedrock surfaces.

The meteorite flux

The marked difference in crater density between the maria and highland surfaces long suggested that the flux of meteorites had changed through time and many early writers had postulated an early intense bombardment of the moon (Urey, 1952; Kuiper, 1954; Kuiper et al. 1966; Hartmann, 1966). Whilst their arguments seemed reasonable it was not until the Apollo landings in the late 1960s and early 1970s that it was possible to date the surfaces independently and establish some estimates of the intensity changes. Dating revealed that initially the flux was intense and that it had declined rapidly, at first with a half life of approximately 10^8 year, until approximately 3.3×10^9 years ago, after which it had remained relatively constant or had perhaps increased slightly (Figure 3.3). The flux of meteorites at the lunar surface evolved in three stages (Hartmann, 1970).

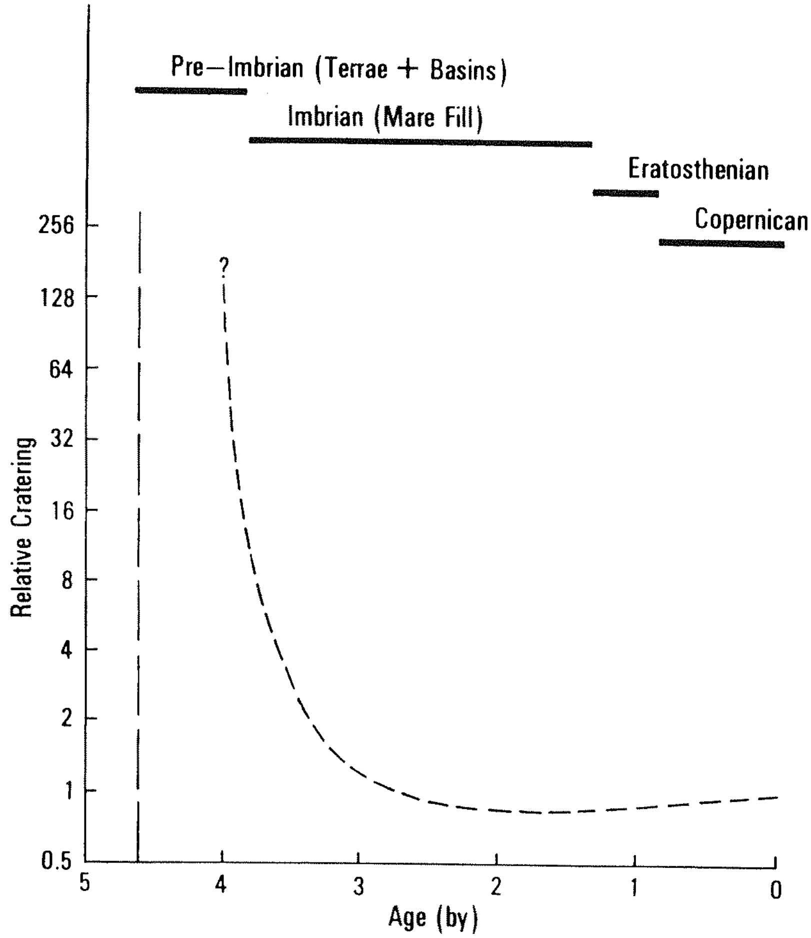

Fig. 3.3. Relative cratering rate (present day flux = 1) of the lunar surface as a function of time (from Hinners, 1971).

(a) An intense bombardment of low velocity (1.7–2 km s^{-1}) circumterrestrial particles left over after the formation of the moon.

(b) An intense bombardment of the last, planet-forming planetesimals swept up from low-eccentricity solar orbits with collision half-lives of the order of 10^8 year. These were medium velocity (2–10 km s^{-1}) collisions which probably formed the large circular maria.

(c) The present phase of high velocity (8–40 km s^{-1}) sporadic meteorites and cometary particles. The modern flux produces approximately 2 to 8 $\times$ 10^3 erg cm^{-2} per year: a very modest energy flux compared to alternative sources, such as wind or water erosion, available at the earth's surface.

Thus, we might expect that soils formed on highland areas differ significantly from those formed on maria. We would expect them to be different compositionally because of the underlying bedrock differences and we could expect that highland soils would be thicker and more highly evolved because of the longer exposure to the meteorite flux and because they were exposed to a more intense flux early in their history (Figure 3.1).

Lunar soil

The surface of the moon is blanketed by a thin layer of weakly-cohesive, low-velocity, detrital material that is generally referred to as "soil" or "regolith" (Figures 3.4 and 3.5; Lindsay, 1976). The use of the term "soil" in a setting where biologic ac-

Fig. 3.4. A clearly defined footprint in the lunar soil at the Apollo 11 site (NASA AS11-40-5878).

Fig. 3.5. Sampling the lunar soil at the Apollo 12 landing site with a drive tube (NASA AS12-49-7286).

tivity is absent is perhaps an extension of the normal usage. However, the earth and its biosphere appear to be unique, at least in the solar system, and there is a need to broaden the use of the term in the extraterrestrial environment. The lunar soil is a continually evolving blanket produced in large part by hypervelocity meteoroid impact (Figure 3.6), although locally a small pyroclastic volcanic contribution may be present. The morphology of small craters penetrating the soil showed that median soil thickness varied from 3.3 to 16 m (Oberbeck and Quaide, 1968). Further, the median thickness of the regolith correlates directly with the density of impact craters and with the age of the underlying substrate (Table 3.1).

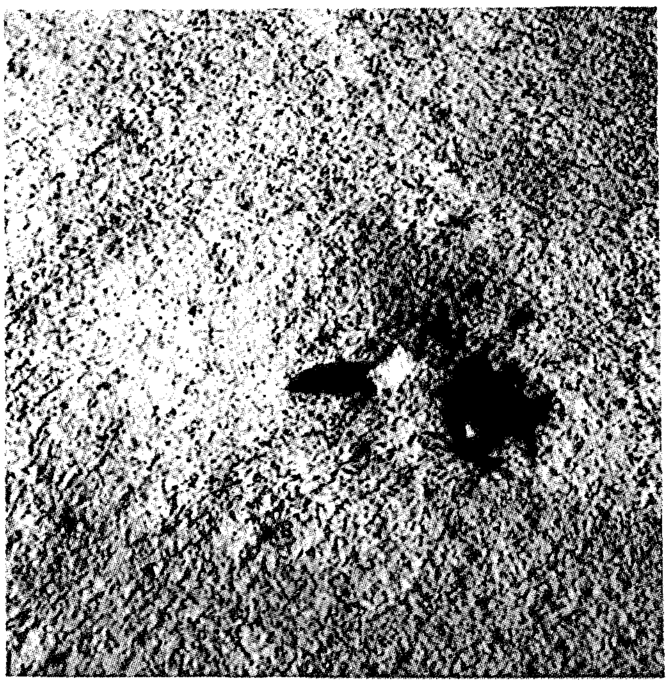

Fig. 3.6. A closeup view of a small crater in the lunar soil surrounded by a thin ejecta blanket. The crater has not penetrated the soil layer and has therefore not added new material to the soil but reworked the existing soil to produced a new stratigraphic unit which may or may not survive future reworking (NASA AS15-85-11466).

TABLE 3.1

Soil parameters at the Apollo landing sites

Apollo site	Soil thickness (m)	Mean grain size (μm)	Age of substrate (b.y.)
11	4.4	15.7	3.51
12	3.7	54.4	3.16
14	8.5	37.2	3.95
15	4.4	28.4	3.26
16	12.2	–	–
17	4.0, 8.0	–	3.74

Soil structure

The lunar soil is not homogeneous but is complexly layered (Lindsay, 1975, 1976). The thickness–frequency distribution of the units is bimodal with the strongest mode at 1.0 to 1.5 cm and a secondary mode at 4.5 to 5.0 cm. The units are distinguished on the basis of colour and texture, particularly variations in grain size. The contacts of the units are generally sharp, although at times difficult to observe because the textural differences are subtle. Some units have transitional contacts and structures similar to flame structures were observed at the base of other units. Both normal and reverse graded beds have been described (Lindsay et al., 1971; Duke and Nagel, 1975).

Physical properties

The mean density of the particulate material forming the lunar soil ranges from 2.90 to 3.24 g cm^{-3} depending upon the nature of source materials. Soils formed on the basaltic maria tend to be denser than soils formed on the more anorthositic highland, that is, the soils in large part reflect the composition of their substrate. The porosity of lunar soils may range from 41 to 70% with the result that soil density is variable (Carrier et al., 1974). The main reason for the range in porosities is probably connected with the abundance of extremely irregularly-shaped glass particles (agglutinates) which are discussed in a following section. Soil density also increases in a logarithmic manner with depth (Carrier et al., 1974). The density profile can be approximated by:

$$p = p_o + 0.121 \ln(Z + 1)$$

where p_o and p are the density at the surface and at some depth Z (cm). p_o is approximately 1.38 g cm^{-3}. The continual reworking of the lunar surface apparently keeps the surface layers of the soil loose but at depth the vibration due to the passage of numerous shock waves causes the soil to increase in density.

Composition

It has been estimated that 95% of a soil sample at any given point on the moon is derived from within 100 km (Shoemaker et al., 1970). The proportion of exotic components decreases exponentially with distance. Thus, the chemistry and mineralogy of the lunar soils, for the most part, reflect the composition of the underlying bedrock (Table 3.2), so that soils from maria areas have an overall basaltic composition with a high Fe content, while the highland soils tend to be more anorthositic in composition and have high Al and Ca values (Table 3.3). The boundaries between the two are sharp at the available resolution (Adler et al., 1972a, b). Lateral movement of detrital material thus cannot be rapid despite continued reworking for long time periods by the meteoroid flux.

TABLE 3.2

Major element chemistry of average Apollo 15 soils and basalts

Basalt	Average soil	Olivine basalt	Quartz
SiO_2	46.61	44.20	48.80
TiO_2	1.36	2.26	1.46
Al_2O_3	17.18	8.48	9.30
FeO	11.62	22.50	18.60
MgO	10.46	11.20	9.46
CaO	11.64	9.45	10.80
Na_2O	0.46	0.24	0.26
K_2O	0.20	0.03	0.03
P_2O_5	0.19	0.06	0.03
MnO	0.16	0.29	0.27
Cr_2O_3	0.25	0.70	0.66
Total	100.00	99.41	99.67

Note: Apollo landed on the lunar maria but at a site close to the highlands (after Taylor, 1975). The higher Al_2O_3 values in the soil reflect the addition of highland materials to the soil.

TABLE 3.3

Average major element chemistry for highland and maria soils

Element	Percent of atoms			Weight percent oxides		
	Mare	Highland	Average	Mare	Highland	Average
O	60.3	61.1	60.9	–	–	–
Na	0.4	0.4	0.4	0.6	0.6	0.6
Mg	5.1	4.0	4.2	9.2	7.5	7.8
Al	6.5	10.1	9.4	14.9	24.0	22.2
Si	16.9	16.3	16.4	45.4	45.5	45.5
Ca	4.7	6.1	5.8	11.8	15.9	15.0
Ti	1.1	0.2	0.3	3.9	0.6	1.3
Fe	4.4	1.8	2.3	14.1	5.9	7.5

Source: Turkevich (1973).

Petrography

The lunar soil consists of three basic components: (a) rock fragments, (b) mineral grains, and (c) glass particles. The composition of these components varies considerably from one locality to another, depending upon the nature of the bedrock. They also vary in abundance in a more local sense (laterally and with depth) in response to the addition of more distant exotic components, the degree to which the soil has been reworked (its age), and in response to bedrock inhomogeneities.

Rock fragments are the dominant soil component in particle size ranges larger than 1 mm (60 to 70% of the 1–2 mm size range). Below 1 mm their importance decreases rapidly as they are disaggregated to form mineral and glass particles. The coarser the mean grain size of the soil the greater, in general, the abundance of rock fragments.

Three categories of clastic fragments should be considered: igneous rock fragments, crystalline breccias and soil breccias.

Igneous rock clasts. Basalts are by far the most common igneous rock fragment in maria soils whilst in the highlands the igneous fragments come from grabbroic-noritic-anorthositic sources or, less frequently, from a highland basalt source. Because the grain size of the minerals forming the highland lithologies is large compared to the size of the soils, positive identification of many rock types is difficult.

Crystalline breccia fragments. These fragments dominate the lithic clasts in all of the highland soils. They are bedrock fragments derived from ejecta units resulting from large meteorite impacts. Fragments of all of the matrix types are represented but again, because of the size of the clasts in relationship to the grain size of the detrital materials forming the breccias, it is not possible to evaluate the relative abundance of different lithologic types.

Soil breccias. The breccia fragments are poorly sorted agglomerations of rock, mineral and glass fragments which to a large extent preserve the texture of the parent rock. They are characterized by a very open discontinuous framework and are estimated to have a void space of about 35%. The breccias have densities of around 2 g cm^{-3} which is only slightly greater than unconsolidated soil (Waters et al., 1971; McKay et al., 1970). The spaces between the larger clasts of the framework are filled with glass-rich clastic materials with an average grain size of about 50 μm (close to the mean grain size of typical lunar soils). The glass forming this matrix tends to be well sorted, closely fitted and plastically moulded against the larger clastic rock fragments. In a general way the texture of these rocks resembles terrestrial ignimbrites (Waters et al., 1971). Glass fragments exhibit a wide variety of devitrification features, ranging from incipient to complete devitrification.

There are two prominent textural features of the soil breccias which provide insights into their depositional history: layering and accretionary lapilli. Very weakly-defined layering or bedding is seen in many of the soil breccias, generally in samples where accretionary lapilli are abundant (Waters et al., 1971; Lindsay, 1972a). Accretionary lapilli appear to be an intrinsic characteristic of the lithology (McKay et al., 1970, 1971; Lindsay, 1972a, b). They are similar to terrestrial volcanic accretionary lapilli (Moore and Peck, 1962) and range in size from 50 μm to 4 mm. The lapilli generally have a core of one or more larger detrital grains surrounded by alternating layers of dark fine-grained glassy material and of larger detrital grains. Soil breccias are thus formed directly from the lunar soil during meteorite impact and are important in the evolution of the soil because they consolidate finer grained materials forming larger particles.

Mineral grains are the dominant detrital particles in the intermediate grain sizes, particularly between 30 and 60 μm, on the coarse side of the mean grain size of the

TABLE 3.4

Modal composition of a mature and an immature highland soil from the Apollo 14 site

Detrital component	Immature soil				Mature soil			
	150–200 (μm)	90–150	60–75	20–30	150–250 (μm)	90–150	60–75	20–30
Agglutinates	5.3	5.2	6.5	12.5	54.5	60.3	56.5	43.5
Breccias	64.4	54.0	53.0	20.7	23.6	23.5	17.5	7.5
Angular glass	2.9	8.8	5.5	12.0	9.4	8.3	11.5	10.0
Rotational glass	0.8	2.0	0.5	2.0	2.4	0.6	2.0	7.0
Mineral grains	11.4	17.0	31.0	50.0	8.8	5.9	8.5	34.0
Rock fragments	15.1	12.4	3.5	2.5	1.0	1.3	1.0	0.0

Note: The mature soil is finer grained and contains an abundance of agglutinates (from McKay et al., 1972).

bulk soil (Table 3.4). This distribution probably reflects the grain size of the source rocks to a large extent, although the physical properties of the minerals themselves are also important. Some minerals, notably feldspar, tend to comminute faster than others.

As is the case for lithic fragments, the detrital mineral grains present in a soil reflect, for the most part, the nature of the underlying bedrock. The grains are generally angular except where rounded grains have been inherited from other source materials. Most mineral grains, but particularly the plagioclase, show much evidence of shock modification as a consequence of the hypervelocity impact environment. Many plagioclase crystals have been disrupted by shock to such an extent that they have been converted to diaplectic glasses.

Plagioclase is the ubiquitous mineral in the lunar soils. The anorthite content is generally high in all soils although the range of values in maria soils is generally greater than in highland soils (Apollo Soil Survey, 1974). A small number of potassic-feldspar grains have been encountered in the lunar soils. Most are small and attached to, or intergrown with, plagioclase.

Pyroxenes are present in most soils and like the plagioclase frequently show signs of shock modification. In maria soils they are almost exclusively clinopyroxene (augite) derived from the maria basalts. Highland soils contain a relatively large proportion (over a third) of orthopyroxene as well as clinopyroxene. The pyroxenes in the highland soils probably were mainly derived from norite and highland basalt clasts in the crystalline breccia basement rock, although pyroxenes are common as mineral clasts in many highland breccias.

The pyroxene/plagioclase ratio of most texturally mature soils consistently increases with decreasing grain size (Table 3.4). This may simply reflect the differing mechanical properties of the two minerals (Lindsay, 1976), or it may be due to fine grain sizes being transported over greater distances (Finkelman, 1973). However, the net result is an increase in mafic elements in finer grain sizes which has considerable bearing on the chemistry of some impact generated glass particles.

Olivine is present in most soils but is very variable in proportion. Only some maria basalts contain a large percentage of olivine so there is a tendency for it to vary from sample to sample at any one mare site. Olivine is generally only present in small amounts (7%) in highland soils, where it is probably derived from pre-existing troctolite fragments and perhaps from a dunite source. In general, the compositional range of olivines is larger in highland soils than in mare soils.

Most soils contain some opaque minerals, mainly ilmenite. The ilmenite is probably largely basaltic in origin at both mare and highland sites. It is very variable in amount and can be quite abundant in some mare soils. A variety of other minerals, such as spinels, is present in the soils but in minor amounts only.

Metallic grains are present in very small numbers in all of the lunar soils. The particles are generally small, few being larger than 100 μm, and consist largely of kamacite and taenite and in some cases schreibersite and troilite. Some particles are free standing metal whereas others are associated with a silicate assemblage. Bulk chemistry of the particles indicates that some come from igneous rocks, especially basalts while others are from meteoritic materials.

Glass particles are abundant in the lunar soils and provide some of the clearest insights into its provenance and to some extent its evolution. Compositionally and morphologically the glasses are extremely complex but they can conveniently be divided into two broad categories: glasses which are essentially homogeneous, and agglutinates which are extremely inhomogeneous.

Homogeneous glasses. The homogeneous glasses are morphologically diverse. However, most are angular jagged fragments obviously derived from larger glass fragments. A smaller number of glass particles have rotational forms (dumbbells, spheres, spheroids (Figure 3.7) and teardrop shapes). Some particles have detrital rock or mineral fragments as cores. Others have a small number of mineral fragments dispersed through otherwise homogeneous glass. Some glasses are vesicular with vesicles forming as much as 30% of the particle volume. In the extreme some spheres are actually hollow bubbles up to 2 cm in diameter. Variations of the morphology of rotational glass forms may be explained in terms of the surface tension of the molten glass, and the angular velocity of the spinning glass mass as it is ejected from the impact crater (Fulchignoni et al., 1971). The regular form of many homogeneous glasses suggests that they are impact melts sprayed onto the lunar surface during crater excavation.

Homogeneous glasses come in a wide variety of colors (colorless, white, yellow, green, orange, red, brown and black) although most particles have darker colors, generally brown or black. In general terms, the darker glasses contain more Fe and Ti whereas the lighter glasses tend to be more aluminous. The lighter colored glasses tend to be more anorthositic or "highland" in composition, whereas the darker glasses tend to be basaltic or "mare" in composition. The angular and rotational glass forms have the same range of colors and refractive indices, suggesting a common origin.

The chemistry of the glass particles generally reflects the composition of local bedrock (Table 3.5). Since the Apollo landing sites were chosen to give insights into the origin of various major features visible on the lunar surface, especially the distinction between maria and highlands, it is instructive to look at chemistry of soils collected at two of these sites, Apollo 14 and Apollo 15. The Apollo 15 site is essentially a mare site, and the Apollo 14 site is situated in the lunar highlands. The Apollo 15 data are particularly instructive in that the site is on the mare surface but a short distance from the highlands (Apennine Front). Both highland and maria rock compositions are, therefore, well represented. To add to the complexity of the Apollo 15 site the soils also contain a modest proportion of green-glass particles of basaltic composition which are believed to be pyroclastic in origin.

A total of eleven compositional types occur among the homogeneous glasses from the Apollo 15 site (Reid et al., 1972), of which the green glass particles are the most common. The green glasses, which are mostly spheres containing a few olivine needles, appear to be pyroclastic materials formed in fire fountains during the flooding of the lunar maria. Similar orange glasses occur at the Apollo 17 site. There is considerable variation in the pyroclastic contribution in surface samples and rotational pyroclastic glass forms occur vertically through the soil blanket.

The soils from the Apollo 14 site contain a similar range of compositional types to the Apollo 15 samples, but the relative proportions of the glasses change in such a way that the amount of highland-derived Fra Mauro basalts increases whereas

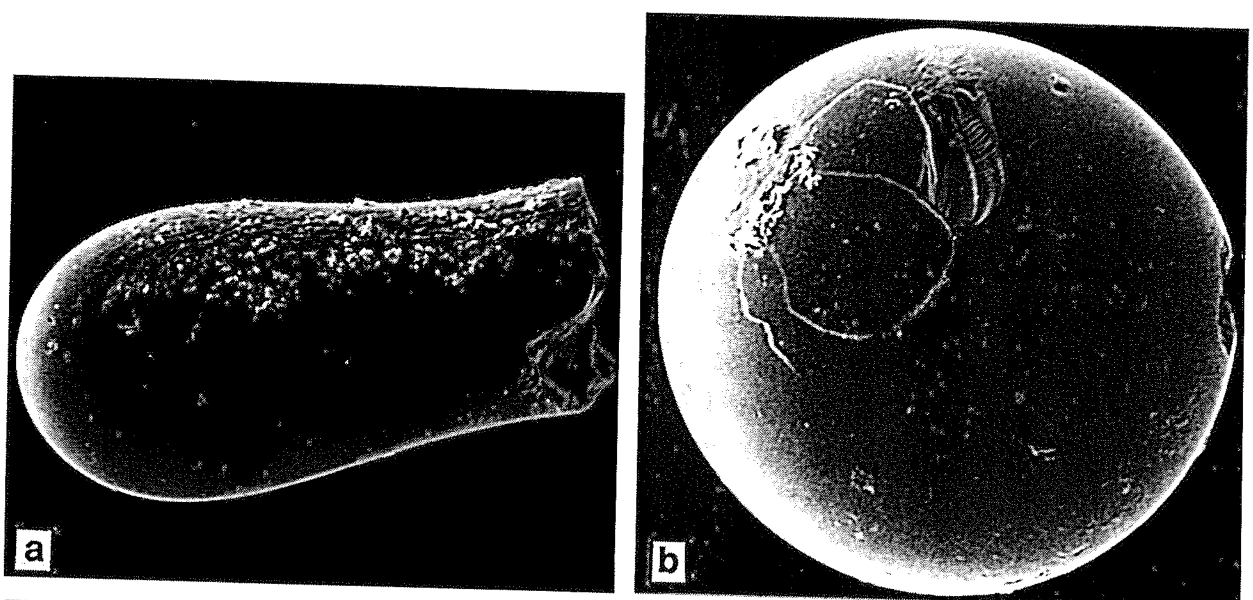

Fig. 3.7. a. A broken homogeneous glass particle with a rotational shape from the lunar soil. The particle is approximately 220 μm in length (NASA S-72-52308). b. An almost perfect glass sphere 390 μm in diameter formed as a response to surface tension in the airless lunar environment. The particle is a basaltic green glass that is probably volcanic in origin but its surface has been spalled by later micrometeorite impacts at the surface of the lunar soil (NASA S-72-53599).

TABLE 3.5

Average composition of the most common homogeneous glass types found at the Apollo 14 site

	Mare type basaltic	Fra Mauro type basaltic	Anorthositic gabbroic	Gabbroic anorthositic	Granitic glass	Low-silica glass
SiO_2	45.48	48.01	45.23	47.37	71.54	37.97
TiO_2	2.77	2.02	0.36	0.14	0.39	0.23
Al_2O_3	10.86	17.12	25.59	31.32	14.15	34.54
FeO	18.14	10.56	5.59	2.98	1.79	1.19
MgO	11.21	8.72	7.84	2.18	0.70	5.57
CaO	9.56	10.77	14.79	14.78	1.97	20.39
Na_2O	0.39	0.71	0.25	0.95	0.93	0.00
K_2O	0.32	0.55	0.12	0.22	6.53	0.00
Total	98.73	98.46	99.77	99.94	98.00	99.89

Source: Apollo Soil Survey (1974).

the maria basalts are of less importance (Table 3.5). A large number of glasses at the Apollo 14 and 15 sites are characterized by a high Fe and low Al content and resemble the composition of maria basalts. Brown ropy basaltic glasses analogous to terrestrial tholeiitic basalts with a higher potassium, rare-earth element and phosphorus content (Meyer et al., 1971) have been found in soils from several Apollo sites. Glasses characterized by a high Al_2O_3 content have been called highland basalt glasses (anorthositic gabbro). This is the anorthositic component generally held to be characteristic of the highlands.

Agglutinates. Agglutinates are one of the most distinctive particle types found in lunar soils and are very important in understanding the origin and evolution of the soil (Lindsay, 1971, 1976). They are intimate mixtures of inhomogeneous dark-brown to black glass and mineral grains, many of which are partially vitrified (Figure 3.8). Compositionally an agglutinate consists of approximately 50% mineral grains. A few agglutinates are vesicular (particularly larger particles, Figure 3.8), most are massive and overall less vesicular than the homogeneous glasses.

Agglutinate surfaces have a coating of fine detrital fragments which gives them a dull saccharoidal texture (Figure 3.8). Most particles are extremely complex in shape but have a general rounded form suggesting that their final shape was determined by the viscosity and surface tension of the fluid glass. Some consist of dendritic glass projections radiating from a central mineral grain while others are bowl shaped or take the form of rings or donuts, suggesting that they are "pools" of melt formed in the bottom of microcraters in fine-grained lunar soil (Lindsay, 1971, 1972c, 1975).

Agglutinates are thus constructional particles made up from fine-grained detrital materials. The included particles are generally finer than 125 μm with a median size of close to 38 μm (Lindsay, 1972c) and a size distribution similar to the fine tail of the bulk soil.

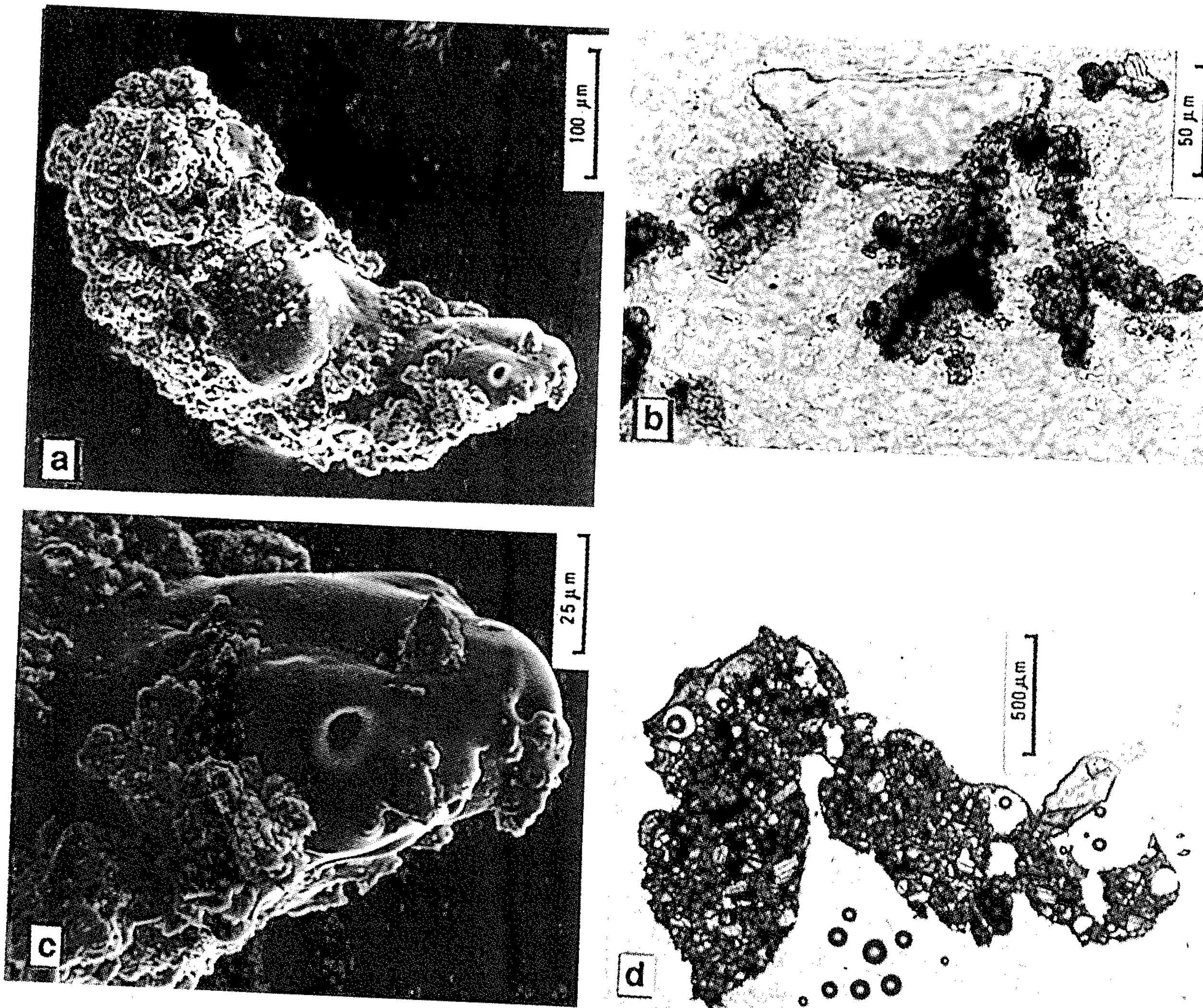

Fig. 3.8. a. A scanning electron microscope image of an agglutinate particle from a mature lunar soil (NASA SP-53160). b. The same particle close up (NASA S-72-53161). c. A very delicate agglutinate consisting of a mineral grain acting as a core with thin projections of brown glass. The maximum length of the particle is 230 µm (NASA S-71-51084). d. A thin section of a large agglutinate (approximately 1.5 mm in length) consisting of fine mineral grains bonded by dark glass (NASA S-71-38420).

Single agglutinates may be as large as 1.5 mm, although rarely, and comminuted agglutinate fragments occur in abundance in particle sizes as small as 16 µm. However, most unbroken agglutinates occur in a narrow size range between about 250 and 178 µm. The proportion of agglutinates in any one sample varies considerably as a function of exposure age (McKay et al., 1974). Their abundance also varies with depth. Typically the size distribution of unbroken agglutinates has a mean of 184 µm and is moderately to moderately-well sorted. The distribution is generally fine skewed, the coarse end of the distribution being truncated. Agglutination thus removes the fine end of the bulk soil grain size distribution and shifts it to the coarse side of the mean grain size of the bulk soil.

The chemical composition of individual agglutinates broadly reflects the composition of the underlying bedrock. Thus, agglutinates from the maria areas have a

TABLE 3.6

Chemical analyses of three agglutinate particles from soils collected at the Apollo 12 site

	Sample 1	Sample 2	Sample 3
SiO_2	46.7	42.2	44.8
TiO_2	2.32	3.34	2.60
Al_2O_3	15.4	9.8	15.6
FeO	12.4	18.3	13.7
MgO	7.9	12.0	7.8
CaO	10.7	8.9	11.6
Na_2O	0.56	0.32	0.54
K_2O	0.92	0.21	0.23
P_2O_5	0.62	0.19	0.17
MnO	0.10	0.12	0.12
Cr_2O_5	0.21	0.35	0.29
NiO	0.00	0.04	0.04
Total	98	96	97

Source: Chao et al. (1970).

general basaltic composition (Table 3.6). Agglutinate glasses may be very homogeneous but in general are very variable in chemistry probably as a result of incomplete mixing (Papike, 1981; des Marais et al., 1973). Because of the inhomogeneity of these glasses more is to be gained from a study of their bulk chemistry than from the chemistry of individual particles. There are significant differences in chemistry between the agglutinate and non-agglutinate fractions of the soils, and neither fraction is comparable chemically to a major homogeneous glass group. In particular it has been found that the composition of agglutinate glasses is very similar to the finest (<20 μm) fraction of the soil (Laul and Papike, 1980). Comminution by impact produces fractionated fine materials because the different mineral components comminute at different rates (Lindsay, 1976; Papike et al., 1982). Fines are then selectively fused by the agglutination process (Lindsay, 1976, Papike et al., 1982) Experimental studies have generally supported these observations (Horz et al., 1984; Simon et al., 1985, 1986).

Texture of the lunar soil

Grain size

Typically the lunar soils are fine grained and poorly sorted. They are moderately coarse skewed and near log normally distributed (Figure 3.9). There is, however, a considerable variability in the grain size parameters of samples from any one site (Lindsay, 1976).

The mean grain size of the soils ranges from a coarse extreme of 380 μm to as fine as 32 μm, although most soils have means close to 62.5 μm. The coarsest soils are believed to be primary detrital materials excavated from the bedrock

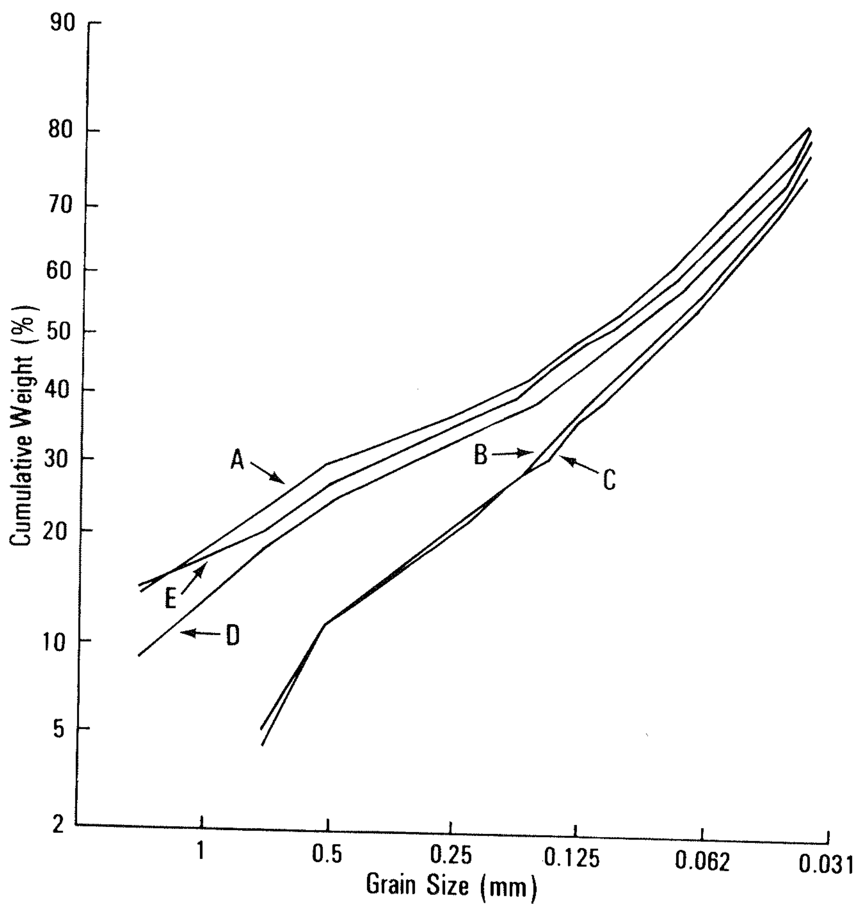

Fig. 3.9. Cumulative grain size distributions for typical lunar soil from the Apollo 11 site (after Lindsay, 1971).

beneath the soil whereas the finest soils appears to be the product of gaseous sorting, perhaps in a base surge. Variations in mean grain size are not random but form part of a time sequence relating to the amount of reworking each soil layer has undergone. When plotted stratigraphically the mean grain size of the soil decreases upwards in a regular manner, with minor erratic excursions which are probably due to the introduction of either older coarse soils or freshly excavated bedrock material (Lindsay, 1973). Further, the mean grain size of the soil is strongly related to the content of agglutinates on the coarse side of the mean (Figure 3.10; McKay et al., 1974). As the soil blanket evolves the accumulation rate decreases and the soil is subjected to longer periods of reworking by micrometeoroids resulting in an increased agglutinate content and a finer mean grain size. It is also apparent from Figure 3.10 that the scatter about the regression line increases as the grain size of the soil decreases. This scatter probably relates to random destruction of large numbers of agglutinates by larger layer-forming impact events.

The sorting of the lunar soils likewise varies. Like the mean grain size, the sorting of the lunar soils conforms to a time sequence in which older soils are better sorted (Lindsay, 1973). There is also a strong relationship between mean grain size and

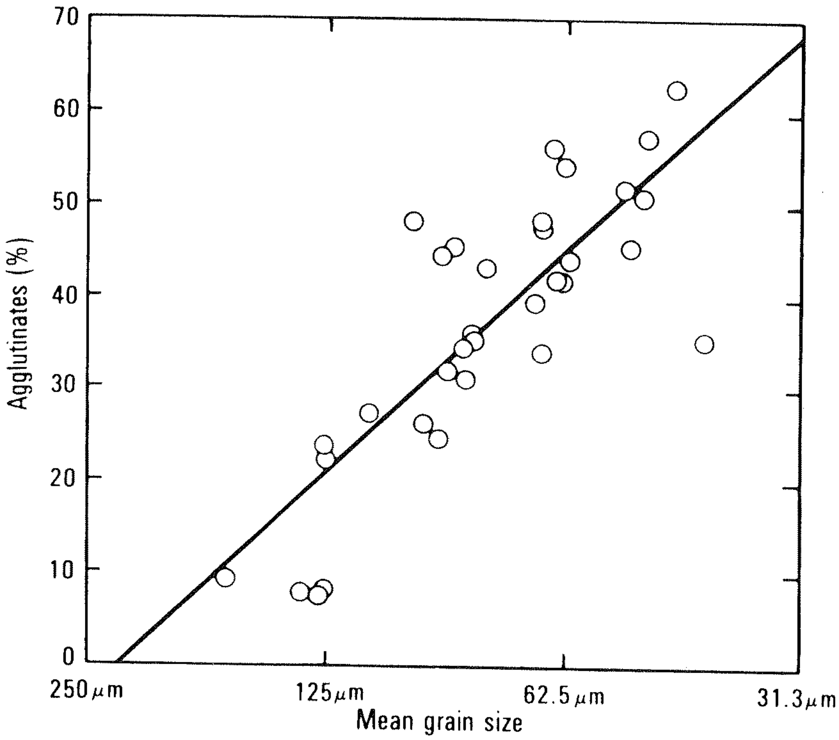

Fig. 3.10. Mean grain size of lunar soils as a function of agglutinate content in the 90 to 150 μm fraction of the soil (after McKay et al., 1974).

sorting which shows that finer soils are better sorted (Lindsay, 1976). Similarly, the better sorted soils contain a greater abundance of agglutinates on the coarse side of the mean.

Agglutinates are extremely delicate, fragile particles that can be destroyed by comminution much more readily than rock or mineral fragments of equivalent size. The relationship between agglutinate content and exposure age indicates that under normal conditions micrometeoroid reworking produces more agglutinates than it destroys by comminution (Lindsay, 1972c, 1976). The destruction of agglutinates reduces the mean grain size of the soil and makes the grain size distribution more symmetrical (that is, less coarse skewed). The larger the agglutinate content of the soil the more dramatic the effect. Once excavated to the surface micrometeoroid reworking again takes effect and the grain size parameters begin to converge on the ideal evolutionary path. We thus see evidence of a series of random excursions during which agglutinates are first formed at the expense of the fine tail of the grain size distribution by micrometeoroid reworking, and are then crushed and shifted back to the fine tail again. This effect has been referred to as cycling and occurs in the most mature soils.

Shape

The shape of soil particles varies considerably, from the smooth spherical form of glass droplets to the extremely irregular and complex shapes of agglutinates.

Between these two extremes are the blocky angular comminuted rock, mineral and glass fragments. The overall mean sphericity of the soil particles is 0.78 (a value of 1.0 being a perfect sphere). However, perhaps predictably, the sphericity of particles varies markedly with grain size and depth in the lunar soil (Lindsay, 1972c, 1974, 1975).

If we look at the sphericity of detrital particles forming a primitive soil (consisting largely of comminuted rock and mineral fragments) we find that sphericity changes in a regular linear manner with grain size. Smaller particles become increasingly more spherical. If we then investigate a texturally mature soil we find that a sinusoidal distribution is superimposed on the general linear trend. Two zones of depressed sphericity develop; one at between 125 μm and 1 mm, the other at between 8 and 31 μm. The midpoint of these two zones lies close to one standard deviation either side of the mean grain size of the bulk soil. The zones of reduced sphericity coincide with whole agglutinates on the one hand and comminuted agglutinates on the other. The relationship between particle shape and size thus relates entirely to the textural evolution of the soil and in particular to micrometeoroid reworking at the lunar surface.

Evolution of the lunar soil

In spite of the tenuous nature of the energy source and the large inefficiencies involved, the meteoroid flux has, over several aeons, produced an extensive body of soil over most of the lunar surface. This soil body probably contains stratigraphic information covering most of the life time of the solar system and perhaps a record of events of galactic magnitude. Clearly the soil body is dynamic and the processes forming it are ongoing which raises questions about its evolution and how information concerning its origins is preserved in the soil stratigraphy.

Energy partitioning and the flux of detrital materials

Lunar soils are the product of hypervelocity impacts that occur when meteorites traveling at orbital velocities strike the planetary surface. The energy released by hypervelocity impact at the lunar surface is partitioned in a complex way. The kinetic energy of an individual impact event is released through heating and either fusing or vaporizing both target and projectile, comminuting the substrate and finally ejecting the heated and comminuted materials. The kinetic energy of the meteorite flux is also partitioned in different ways according to the mass spectrum of that flux. Different portions of the mass spectrum contribute to, or modify, the soils in different ways (Figure 3.11). It is as yet difficult to evaluate these variables but some order of magnitude estimates have been made (Lindsay, 1975). The post-mare meteoroid flux at the lunar surface produces a primary sediment flux of the order of 2.78×10^{-7} g cm^{-2} per year, which represents an erosional efficiency of approximately 1.6% (Gault et al., 1972; Lindsay, 1975). The sediment flux per unit

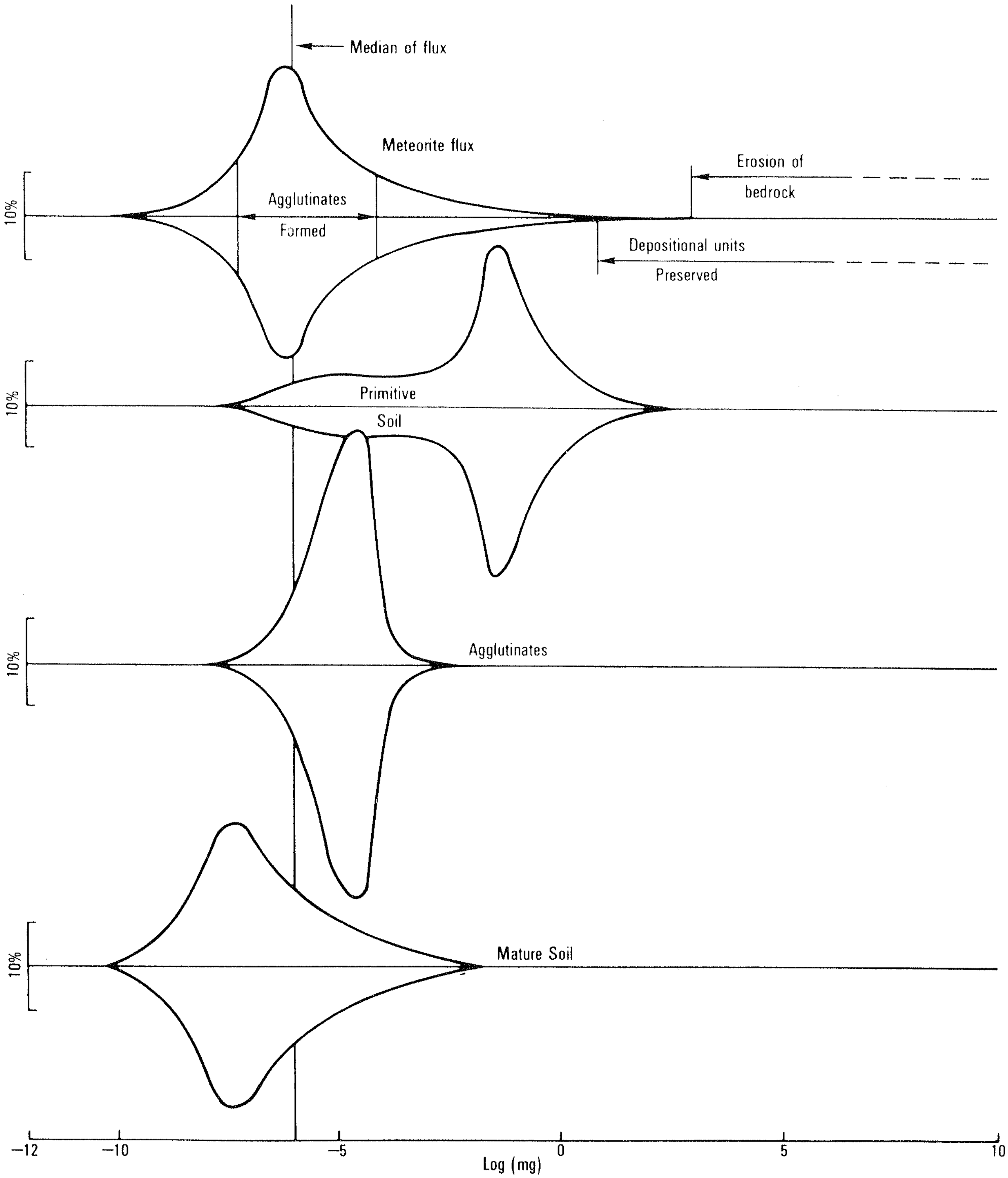

Fig. 3.11. The size distribution of the meteoroid flux, a primitive lunar soil, a mature lunar soil and the agglutinates from a mature lunar soil all expressed on the same scale as weight percent per decade of mass to allow direct comparison (from Lindsay, 1975, 1976).

area entering the earth's oceans is 175 times larger than the lunar sediment flux. Bedrock material appears to be excavated most efficiently by meteoroids larger than 10^3 to 10^4 g. In contrast, agglutination is caused by micrometeoroids in the mass range of 10^{-7} to 10^{-4} g or by about 68% of the flux mass. Layer-forming events appear to be the product of meteoroids larger than 7 g or less than 1% of the flux mass.

Soil accumulation is a self-damping process such that the average accumulation rate decreases with time (Lindsay, 1976; Quaide and Oberbeck, 1975). If the meteoroid flux had remained constant over time its effectiveness as an agent of erosion would gradually be reduced as the soil blanket grew in thickness (Figure 3.12). For new material to be excavated from the bedrock beneath the soil an impact must be energetic enough to penetrate the pre-existing soil layer. As the soil blanket grows, more and more energetic events are required to accomplish the same result. However, it is evident that the number flux of particles decreases rapidly with increasing particle size and with it the available erosional energy must also decrease. The energy actually available for erosion of bedrock is probably considerably less than 1% of the total meteoritic energy incident on the moon (Figure 3.11; Lindsay, 1975). Whatever the history of bombardment, there should be rapid initial accumulation of soil followed by gradually decreasing growth rates.

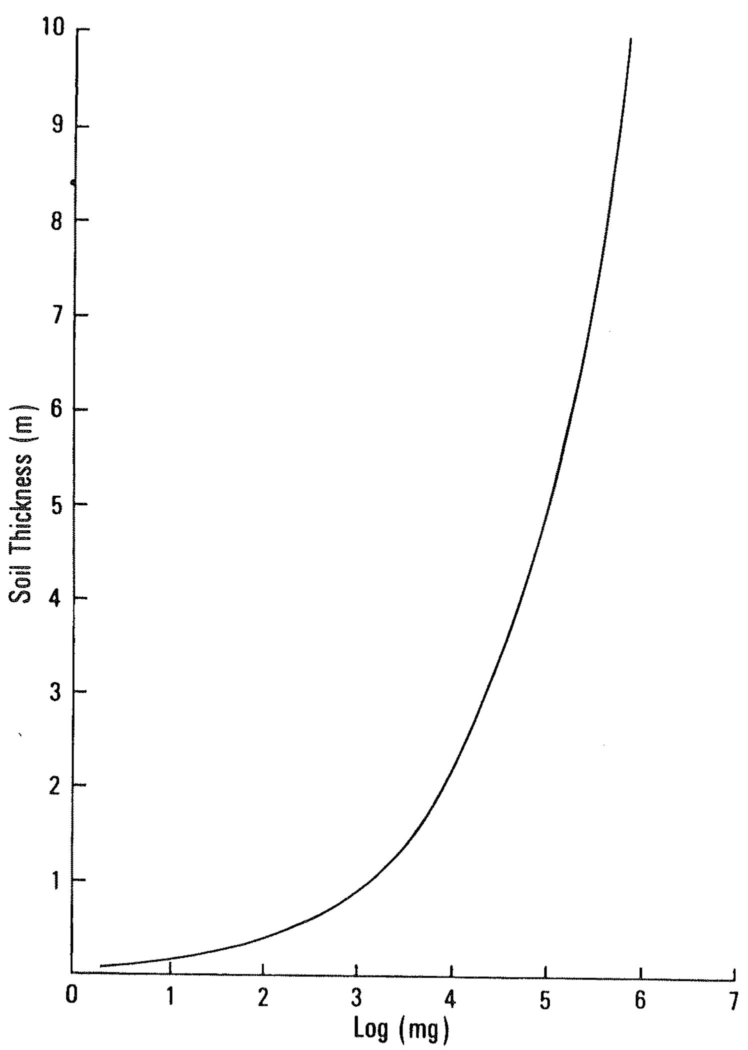

Fig. 3.12. Soil thickness in relation to the minimum meteoroid mass capable of penetrating the soil to excavate bedrock (after Lindsay, 1975).

The variations in median soil-thickness thus reflect differences in elapsed time since the production of the new rock surface upon which the soil is evolving. The soil begins as a thin deposit of nearly uniform thickness and gradually changes to a thicker deposit with a greater spread of thickness values. That is, the standard deviation of the soil thickness frequency distribution increases with time. The increasing spread of thickness values highlights the stochastic nature of the impact process — soil accumulation is a discontinuous process dependent upon random (in both space and time) hypervelocity impacts.

Modelling of such complex processes requires more knowledge than is currently available about the meteoroid flux, and therefore some simplifying assumptions are necessary. Quaide and Oberbeck (1975) have used a Monte Carlo approach to study soil development on maria surfaces. If the meteoroid flux has been constant since the flooding of the maria the thickness (Th) of the soil blanket is related to its age (A) by:

$$Th = 2.08A^{0.64}$$

where Th and A are in meters and aeons, respectively. The accumulation rate (dTh/dA) decreases exponentially with time. This observation has considerable bearing on the understanding of the texture and mineralogy of the soil. As the accumulation rate declines more and more of the kinetic energy of the meteoroid flux is redirected into mixing and reworking the soil. This extra energy is particularly important in modifying the grain size parameters of the soil which in turn affects the nature of the impact-produced glass particles and the glass content of the soil. Because smaller volumes of fresh bedrock material are added with time, any materials that are cumulative will increase in concentration; this is true of impact glasses (Lindsay, 1971, 1972c) and should be true of the meteoritic materials themselves.

Textural evolution

The texture of the lunar soil evolves in a regular manner in a direct response to continued reworking by the meteoroid flux (Lindsay, 1976). The two main dynamic processes responsible for this developmental sequence are comminution and agglutination. The textural maturity of the lunar soil is determined almost entirely by the balance between these two opposing processes, one destructive the other constructive. As a result of this complex interaction between the soil and the meteoroid flux the soil passes through three transitional evolutionary stages (Lindsay, 1975). For convenience the stages are designated: the comminution dominated stage, the agglutination dominated stage and the steady state or cycling stage (Figure 3.13).

The comminution dominated stage

Because of the extreme age of the lunar surface few soils have survived that could be assigned to this evolutionary stage. Energetic impact events striking lunar

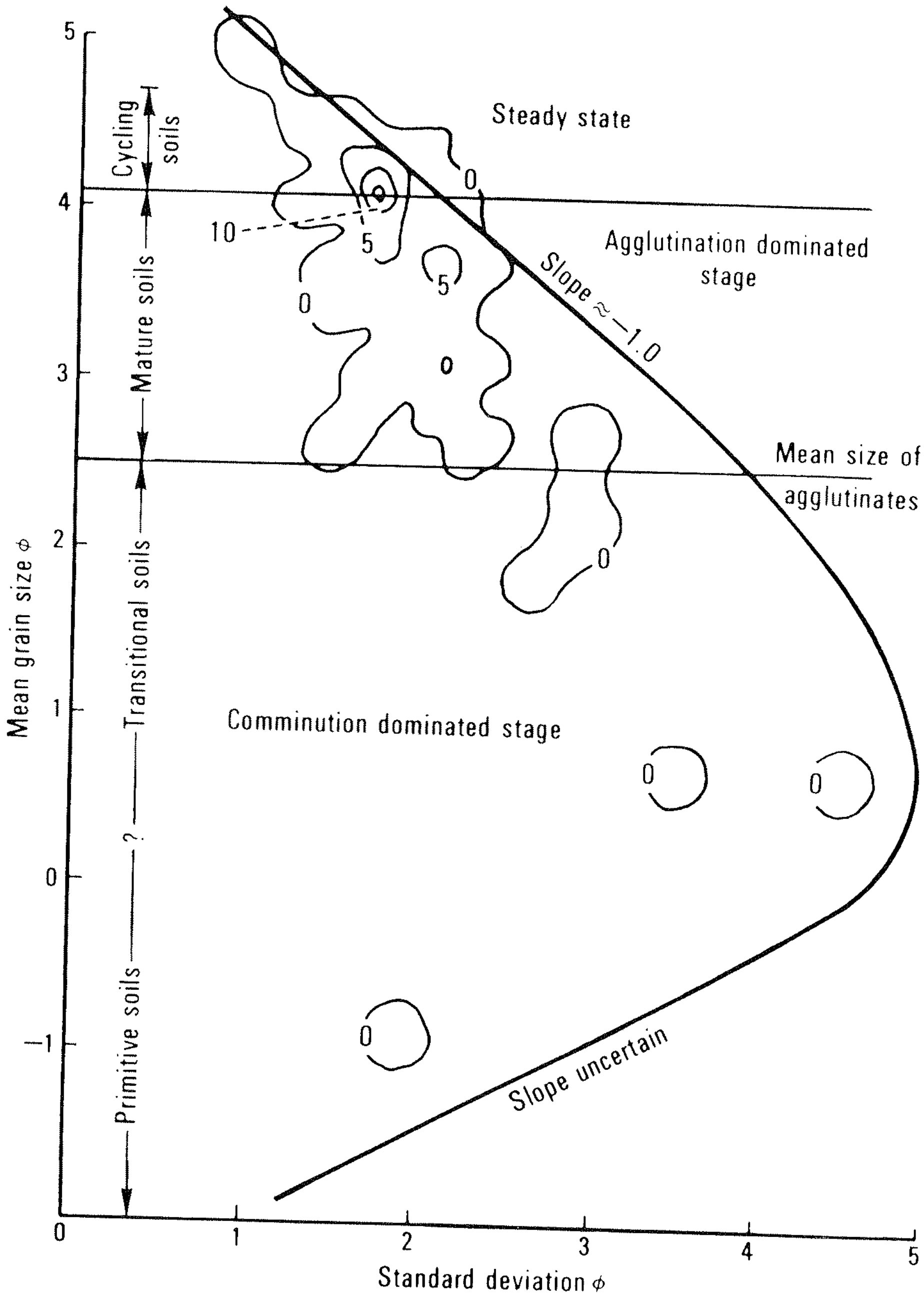

Fig. 3.13. Stages in the evolution of the lunar soil shown by changes in mean grain size and sorting. The solid line indicates the evolutionary path of an idealized soil. Contours show the density distribution of analyzed soil samples. Note that most soils are finer than the mean agglutinate size and they concentrate at a mean of 58 μm (4.1 phi), the stable point between the two glass modes. Grain size is expressed in phi units where grain size in mm = $2^{-\text{phi}}$.

bedrock produce coarse grained and poorly sorted ejecta. When such loose particulate materials are subjected to repetitive fracture the grain size distribution is gradually modified in such a way that it asymptotically approaches a log-normal form (Kolmogoroff, 1941; Halmos, 1944; Epstein, 1947). The mean grain size of the particulate materials is gradually reduced and they become more poorly sorted. These primitive soils are coarse grained with means in excess of 1 mm. During this stage of soil development comminution is the only effective process as most meteoroids interact with single detrital particles. The particles of soil are so large that most micrometeoroid impacts are either involved in catastrophic rupture of grains or they simply form pits on the larger grains.

Experimental studies suggest that cratering occurs in loose particulate material only when the particle mass is less than an order of magnitude larger than the impacting projectile. Thus comminution will dominate soil dynamics until the soil particles are reduced to masses of the order of 10^{-5} g, that is, an order of magnitude larger than the mean mass of the meteoroid flux (10^{-6} g). This suggests that the soils would have a mean grain size of the order of 95 μm. At the very least it is unlikely that agglutination can become effective until the mean grain size of the soil is less than 125 μm, the mean grain size of the agglutinate distribution.

The agglutination dominated stage

Most of the soils presently at the lunar surface appear to be passing through this stage of development. Once the mean grain size of the soil is small enough for agglutination to be effective, it begins to balance the effects of comminution (Lindsay,1976). This results in a continuing decrease in the mean grain size as larger lithic and mineral grains are comminuted; the standard deviation of the grain size distribution is reduced causing the ideal evolutionary path in Figure 3.13 to reverse its slope. During this stage there is a strong correlation between mean grain size and standard deviation, and the distribution becomes skewed towards the coarser grain sizes. The glass particles produced by the two opposing processes (agglutination and comminution) are formed in two narrow size-ranges. Agglutinates concentrate at approximately 177 μm while comminuted agglutinates accumulate at approximately 16 μm. The mean grain size of most lunar soils lies between these two glass modes (Lindsay, 1972c). Ultimately, as the coarse rock and mineral fragments are broken down by comminution, the mean grain size is forced to lie between the two glass modes at close to 63 μm and the standard deviation of the soils is roughly coincident with the glass modes. At this point the soils are approaching a steady state. However, agglutination still dominates in that more agglutinates are formed than are destroyed by comminution.

The grain size distribution of the agglutinates appears to be controlled by a complex of variables, but to a large extent it is a direct function of the meteoroid mass–frequency distribution. At the root-mean-square velocity of 20 km s^{-1} a meteoroid fuses or vaporizes about 7.5 times its own mass of detrital materials (Gault et al., 1972; Lindsay, 1975). Experimental work shows that most agglutinates are formed at shock induced pressures of between 381 and 514 kbar (Gibbons et al., 1975). Assuming that the detrital material enclosed in the agglutinate forms part of the solid–liquid-mixture transition, an agglutinate should be about 5 times the mass of the impacting meteoroid. The size distribution of agglutinates on the fine side of 217 μm fits the model well. However, on the coarse side of 212 μm the size distribution is depleted by comparison with the micrometeoroid flux. This implies that agglutination is inhibited by a second variable, which relates to scale (Lindsay, 1975). The most likely variables are viscosity and surface tension of the melt. Agglutinate formation relies on the ability of the melt to maintain its integrity during crater excavation. As event magnitude increases the probability that an agglutinate will form decreases, because the glass is dissipated as small droplets. Consequently, the agglutinate grain size distribution is fine skewed and by comparison with the meteoroid flux it is excessively deficient in coarser particles. The standard deviation of the agglutinate distribution cannot, therefore, be larger than the standard deviation of the meteoroid flux; it is in fact about half of the equivalent flux value.

The mean agglutinate size distribution suggests that most agglutinates are formed by micrometeoroids in the mass range 5.5×10^{-5} to 7.0×10^{-8} g. This mass range includes approximately 68% of the mass and hence of the kinetic energy of the meteoroid flux. Obviously, agglutinate formation is one of the major processes active in the reworking of mature lunar soils.

The steady state stage

As the evolving soil moves closer to the point of stabilization, where the mean lies between the glass modes, the grain size parameters become more and more dependent on the presence of the agglutinate mode.

At this point a soil may enter a cycling mode. Agglutinates are fragile and readily crushed so that when a soil is redistributed by a larger layer-forming impact event a large proportion of the agglutinate population is destroyed. Removal of the agglutinates causes the mean and standard deviation to decrease and become more coarse skewed. That is, the soil tends to take on the grain size parameters of the second glass-mode. The soil is then exposed at the lunar surface to the micrometeoroid flux and agglutination once again becomes effective, shifting the grain size parameters back to the steady state (Lindsay, 1975).

Probabilistic models for mixing and turnover rates due to hypervelocity impact suggest that the upper millimeter of the lunar soil is the primary mixing zone. This mixing zone is probably coincident with the zone in which agglutinates form. Below this mixing zone the rate of turnover and mixing decreases very rapidly with increasing depth which explains the well-preserved soil stratigraphy. Surface soil layers (Figure 3.6) appear to survive from about 1 to 50 m.y. before burial (Gault et al., 1972).

The presence of normal and reverse graded bedding in the lunar soil suggests that a single set of depositional processes were operative during the formation of at least some stratigraphic units. Grain size data from individual units suggests that two processes may be active in the transport of soil materials following a hypervelocity impact: base surge and grain flow (see Lindsay, 1976).

This model represents an ideal situation where processes are assumed to be operating in a closed system. In reality the soils are continually reworked and soils from two or more stages are continually being mixed. Soils from earlier erosional episodes may be fossilized by burial and later re-excavated to mix with more mature soils at the lunar surface (Lindsay, 1976). At the same time larger impact events may penetrate the soil blanket and excavate bedrock material. The result is a complex interbedded sequence of mixed soils that gradually converge to a steady state upward through the stratigraphy (Figure 3.13).

Soil maturity

As the lunar soils evolve they change in texture and composition, that is, they mature. The term "maturity", when applied to detrital material in a terrestrial setting, is generally used in the sense of a measure of chemical and physical progress along some predetermined evolutionary path. If we consider sand accumulating on a beach we see several processes that operate to modify the texture and composition of the detrital materials. Wave action abrades, rounds and sorts the detrital particles while less stable minerals deteriorate chemically in the aqueous environment and are removed. The end product is a well-sorted sand consisting of well-rounded quartz grains. In the case of beach sand the proportions of detrital quartz can be used as a measure of maturity. Thus in Pettijohn's (1957) concept of maturity monomineralic quartz sand is "the ultimate end product" to which beach sand is "driven by the formative processes that operate upon it".

To apply Pettijohn's (1957) concept to the waterless depositional environment of the lunar soil requires a deeper insight into what we are actually attempting to measure with a maturity index. The answer is simply the total energy expended in the formation of the sediment (Lindsay, 1971, 1976). The formative processes operating on the lunar soil in large part relate to the kinetic energy of the meteoroid flux at the lunar surface. Each impact on the lunar surface releases a significant proportion of its kinetic energy in the form of heat energy (Gault and Heitowit, 1963; Braslau, 1970). In turn a proportion of this heat energy is expended in fusing some of the detrital materials excavated by the impact. Since the soil is continually reworked by the meteoroid flux the glass content of the soil increases with time. Thus the total glass-content of the soil is a measure of mineralogic maturity (see Lindsay, 1976).

However, the maturity index is not a simple linear function with time. While it is obvious that the total glass-content of the soil must increase with time the rate of increase in the glass content will decline exponentially as the accumulation rate decreases and more energy is expended in remelting pre-existing glass. The agglutinate content of the soils has also been suggested as a measure of maturity. This measure offers a more sensitive index of textural maturity during the later stages of soil development and reflects more the recent exposure age than the overall soil maturity (Lindsay, 1976). However, very young or very old soils do not conform to the agglutinate model.

Other worlds

In spite of the obvious differences in mineralogy, chemistry and texture between lunar and terrestrial soils from the view point of soil mechanics and engineering they share many similarities. Because they were formed in a vacuum in an environment free of the effects of oxygen and water the fine soil particles have "clean" surface that allow electrostatic bonding and clumping to occur readily with the result that the lunar soils behave in much the same manner as might be expected of a granular, slightly cohesive (perhaps damp) terrestrial soil. An astronaut stepping onto the surface of the moon for the first time walked on soil with a familiar physical feel, if not a familiar environment. The soil is firm and it retained clear impressions of the astronaut's boots (Figure 3.4).

Soils occur on most if not all planetary surfaces. They are the product of geologic processes and reflect the planet's history (Figure 3.14). Through lack of knowledge it is difficult to know what we might expect on the Jovian planets but on the smaller planets such as Mercury or the Asteroids we can confidently predict a soil very much like that existing on the moon. Similarly, on planets intermediate in size between the earth and the moon such as Mars we see what might best be described as hybrid soils where the meteorite flux has been involved in generating clastic materials that have then been modified and redistributed by the effects of a tenuous but significant atmosphere. Venus likewise seems to share many features in common with the Martian setting for although its atmosphere is dense it does not have a hydrosphere and the resulting surface features suggest a combination of meteorite impact and eolian processes.

At the lower end of the size range, rocky asteroids of modest size (300 km) could be expected to have regoliths as much as 3.5 km thick. Asteroids smaller than 300 km (and small satellites such as those of Mars) will have thin regoliths because of ejecta loss resulting from their small gravitational fields. Unlike the moon the

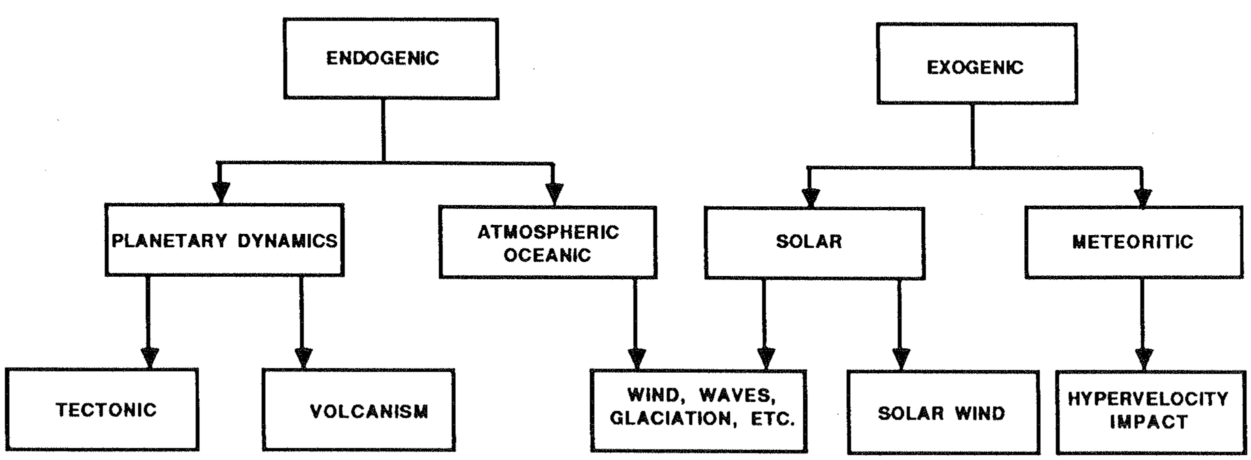

Fig. 3.14. Simplified schematic outlining processes affecting soil formation on planetary surfaces.

regoliths on these small bodies will not show evidence of extensive reworking because the rate at which ejecta blankets form exceeds the excavation rate (Bunch and Rajan, 1988). Evidence of the importance of meteorite impact on the surfaces of the smaller planets and satellites is abundant in the form of intense cratering. All satellite surfaces so far studied, with the exception of Io and to a lesser extent Europa, are cratered, as are Mars and Mercury. In the case of Mercury the moon is probably a very effective analog, whilst on Mars and Venus the atmosphere plays an important role in soil formation, and subdues the role of hypervelocity impact. The spectral characteristics of Mercury suggest that it is covered with a mature lunar-like soil. Io is exceptional, for in spite of its modest size gravitational interaction with the main planet results in active volcanism; soil formation is therefore dominated by endogenic processes. Europa is sheathed in ice and has more in common with the other outer planet satellites which consist of mixtures of water frost and dark silicate or carbonaceous materials. All of the icy satellites are expected to have impact generated regoliths. Evidence of impacts is clear and there are significant variations in albedo that imply reworking of the surface. Endogenic processes were also important on many of these larger satellites. However, "water vulcanism", for example, is little understood. Latitude dependent variations in albedo in some cases suggest other active surface processes, perhaps relating to temperature, but as yet not understood (Veverka et al., 1986). On other less well understood satellites such as Titan and Triton, which have significant atmospheres, other processes come into play; the soils there would be the product of chemical weathering and other atmospheric processes, much as on earth but modified by their distance from the sun and reduced solar radiation. The formation of soil is thus a complex process dependent upon many variables: the size of the planet, the composition of its surface, its distance from the sun and, in the case of satellites, its orbital relationship to the main planetary body (Figure 3.14).

Given all of these complexities, understanding the lunar soil is all the more important because it provides a point of reference for any planetary study. While only a superficial blanket on a planetary surface it is the soil that provides the first point of contact for any scientific study, whether the study be carried out remotely from earthbound telescopes or orbital platforms, or by direct observations using manned or unmanned spacecraft on the planetary surface. Clearly, much remains to be learnt about soils or regoliths on other planetary surfaces. Manned landings on Mars planned for the early 21st century are likely to extend our knowledge of surface processes significantly, but the main progress in understanding extraterrestrial soils is likely to come from unmanned planetary missions. Surface landings by unmanned spacecraft offer the greatest potential, but, in the case of the smaller planets and asteroids lacking an atmosphere, orbital experiments using spectral characteristics and remote methods such as passive X-ray fluorescence will probably provide the great bulk of data for the foreseeable future.

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